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by

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ABSTRACT

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It is possible in principle that the nighttime F layer is maintained either by H^+ ions in the protonosphere providing a source of O^+ ions through charge exchange, or by an upward drift of ionization which raises the layer to an altitude where recombination is slow.

The steady state solution of the H^+ and O^+ ion distributions and fluxes when the H^+ ions act as a source for the F-layer is presented. It is shown that, on the average, the number of H^+ ions that would be consumed by this process cannot be supplied by upward diffusion in the daytime. (A similar conclusion was reached by the authors previously but from a more restricted calculation.)

The alternative hypothesis for maintaining the nighttime F-layer by an upward drift of ionization is examined in some detail. The anaylsis is applied to a particular nighttime ionosonde record which is shown to be consistent with a vertical drift velocity of at least 30 m/sec.

A method for deducing the drift velocity from $h_m F2$ and other ionospheric parameters is presented. It is suggested that this information can be used in conjunction with ionization drift velocity data to uniquely determine the ionospheric electric field and the southward component of the atmospheric wind system at F2 peak heights.

1. INTRODUCTION

It has long been realized that at moderate latitudes the nighttime F layer, particularly in winter, decays much more slowly in the hours before sunrise than during the hours after sunset (Croom, Robbins and Thomas, 1959; Wright, 1962). This behavior can be explained either in terms of a nighttime source of ions, or by a change in the effective recombination coefficient for the layer. Two ways by which the nighttime ionization could arise are from a direct ionization mechanism, e.g., from energetic particles, or by charge exchange of the H^+ ionization reservoir in the protonosphere with atomic oxygen. Changes in the effective recombination coefficient of the F layer could arise from an actual change in the chemical composition of the atmosphere or by a vertical translation of the F layer into a region of lower atmospheric density; only the later mechanism could produce large changes in the actual loss rate. Changes in the rates of chemical processes with temperature cannot be excluded, but little is known of the temperature coefficients for the pertinent reactions.

The possibility that the protonosphere may act as a nighttime ionization source is re-examined in this paper, with an approach which is somewhat more realistic than has previously been made (Hanson and Patterson, 1963). It is found that the protonosphere cannot maintain the nighttime F layer on a regular basis since it is not possible during the day to generate the requisite return flux of protons.

The detailed examination of a particular nighttime ionosonde record which was confirmed with results obtained with a Blue Scout rocket (Sagalyn and Smiddy, 1964) shows that there must have been an upward drift of the F-layer ionization at the time of that measurement. In this particular case, no nighttime ionization source would have been required. A relationship between the height of the F2 peak and the vertical ionization drift velocity is obtained

for the case where the loss coefficient, β, has the scale height of molecular nitrogen.

2.   THE DIFFUSION EQUATIONS

2.1   General Equations

The equations describing the mutual diffusion of singly charged ions and neutral particles in the ionosphere will now be derived. The equations of motion of a multicomponent gas are given by (Chapman and Cowling, 1960),

$$ n_s m_s \frac{D_s c_s}{Dt} = n_s F_s - \nabla p_s + kT_s \sum_j n_s n_j (c_j - c_s)/b_{sj} \quad . \tag{1} $$

D_s/Dt is the time derivative following the motion of gas s; n_s, m_s, c_s, p_s and T_s are respectively the concentration, mass per particle, mean velocity, partial pressure and temperature of gas s, and k is Boltzmann's constant. F_s is the force per particle acting on gas s. It is assumed that all the ions and neutral particles have the same temperature T. The collision parameters b_{ij} are defined according to,

$$ D_{ij} = b_{ij}/\sum_j n_j , \tag{2} $$

where D_{ij} is the binary coefficient of diffusion for gases i and j at the pressure of the entire mixutre. The summations in (1) and (2) are to be taken only over ions and neutral particles so that the resistances encountered by diffusion of the heavy particles through the electron gas are neglected. Further, (1) only applies in an isothermal region where thermal diffusion need not be included. Setting

$$ p_s = n_s kT_s , \tag{3} $$

neglecting the acceleration term on the left hand side of (1) and noting that in the absence of a magnetic field,

$$\underline{F}_s = m_s\underline{g} + e_s\underline{E}, \tag{4}$$

where $\underline{g}$ is the acceleration of gravity, e_s is the electric charge on the particles of gas s, and $\underline{E}$ is the electric field vector, gives,

$$\nabla n_s = \frac{n_s m_s}{kT}\underline{g} + n_s\frac{e_s\underline{E}}{kT} + \sum_j n_s n_j(\underline{c}_j - \underline{c}_s)/b_{sj}. \tag{5}$$

The electron concentration gradient is given by

$$\nabla n_e = -n_e\frac{e\underline{E}}{kT_e}, \tag{6}$$

where T_e is the electron temperature, n_e is the concentration of electrons and e is the magnitude of the charge on the electron. The collision terms and the term containing the electron mass in (5) have been neglected in deriving (6).

The usual assumption is made that the net space charge in the ionosphere is zero. Since only singly charged ions are considered, this requires that

$$n_e = \sum_s{}' n_s, \tag{7}$$

where $\sum{}'$ denotes summation over the ions alone and excludes neutral particles as well as electrons. It is further assumed that the ionosphere is horizontally stratified.

Taking the z-component (vertical) of (5), summing over all the ions, and using (7) and the z-component of (6) gives

$$\frac{eE_z}{kT} = \frac{T_e}{n_e(T+T_e)}\left[\sum_s{}' n_s/H_s - \sum_s{}'\sum_j{}'(n_s\phi_j - n_j\phi_s)/b_{sj}\right], \tag{8}$$

where

$$H_s = kT/m_s g(z) \tag{9}$$

and ϕ_j is the z-component of $n_j \underline{c}_j$. The z-component of (5) is thus

$$\frac{\partial n_s}{\partial z} = -\frac{n_s}{H_s} + \frac{e_s T e n_s}{e n_e (T+T_e)} \left[\sum_s \frac{'n_s}{H_s} - \sum_s \sum_j (n_s \phi_j - n_j \phi_s)/b_{sj} \right] + \sum_j (n_s \phi_j - n_j \phi_s)/b_{sj} \; . \quad (10)$$

In addition each species must satisfy the equation of continuity,

$$\frac{\partial n_s}{\partial t} + \frac{\partial}{\partial z} \phi_s = S_s - L_s, \quad (11)$$

where S_s and L_s are the production and loss rates per unit volume for species s. The presence of a magnetic field which is not vertical changes the simple situation described above. When there are no external electric fields, the component of the forces on the plasma perpendicular to the magnetic field establish a current $\underline{j}$ such that $\underline{j} \times \underline{B}$ counteracts these forces. External electric fields (e.g., the polarization field arising from the dynamo current system) on the other hand cause the plasma to drift with a velocity $\underline{E} \times \underline{B}/B^2$. In effect, the plasma is constrained to follow the magnetic field lines above 200 km where the ion and electron collision frequencies are small compared to their gyrofrequencies. If the neutral gas is at rest and there are no external electric fields, the magnetic field may be taken into account in (10) by multiplying the collision coefficients b_{ij} by $\sin^2 I$, where I is the magnetic dip angle. (For small I more elaborate steps need to be taken.)

In section 4 the diffusion of a plasma containing a single species of positive ions will be discussed. Dougherty (1961) has considered this case in some detail and has taken into account the presence of imposed electric fields and horizontal motions of the neutral atmosphere. His results show that the continuity equation for a horizontally stratified medium becomes modified by the introduction of a vertical drift of the ionization with velocity w, and for atomic

oxygen ions takes the form,

$$\frac{\partial n_3}{\partial t} + \frac{\partial \phi_3}{\partial z} + w \frac{\partial n_3}{\partial z} = S_3 - L_3, \tag{12}$$

where w is the same for the ions and electrons if no net currents flow along the magnetic field lines. w is assumed independent of altitude, and is determined by the imposed electric fields and the horizontal motions of the neutral atmosphere, which is assumed to be independent of z. n_3 is the concentration of atomic oxygen ions and S_3 and L_3 are the production and loss rates of atomic oxygen ions. A further modification should properly be made to (12) to take into account the vertical divergence of the magnetic field lines. The magnitude of this effect has been calculated, however, and its neglect does not affect any of the conclusions of this paper.

2.2 The Steady State Diffusion of H^+, O and O^+

The steady state diffusion of hydrogen ions, oxygen atoms and atomic oxygen ions is of particular interest in this study. Expanding (10) gives

$$\frac{\partial n_1}{\partial z} = - n_1/H_1 + \frac{n_1 T_e}{(n_1+n_3)(T+T_e)} \left[\frac{n_1}{H_1} + \frac{n_3}{H_3} + n_2 \left(\frac{\phi_1}{b_{12}} + \frac{\phi_3}{b_{23}} \right) \right] - n_2\phi_1/b_{12}$$

$$\tag{13}$$

$$- (n_3\phi_1 - n_1\phi_3)/b_{13}$$

$$\frac{\partial n_2}{\partial z} = -n_2/H_2 + n_2\phi_1/b_{12} + n_2\phi_3/b_{23} \tag{14}$$

and

$$\frac{\partial n_3}{\partial z} = - n_3/H_3 + \frac{n_3 T_e}{(n_1+n_3)(T+T_e)} \left[\frac{n_1}{H_1} + \frac{n_3}{H_3} + n_2 \left(\frac{\phi_1}{b_{12}} + \frac{\phi_3}{b_{23}} \right) \right]$$

$$\tag{15}$$

$$- n_2\phi_3/b_{23} + (n_3\phi_1 - n_1\phi_3)/b_{13},$$

where the subscripts 1, 2 and 3 refer respectively to hydrogen ions, oxygen atoms and atomic oxygen ions. It is assumed that ϕ_2, the vertical flux of 0, is zero. The last two terms on the right hand side of (14) are small compared to n_2/H_2 so that atomic oxygen can be taken to be in diffusive equilibrium.
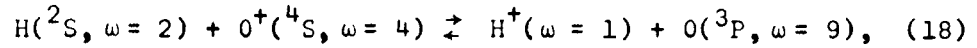
In the steady state situation, the fluxes of H^+ and 0^+ must also satisfy the equations of continuity,

$$\frac{\partial \phi_1}{\partial z} = K_{18}(n_3 n_4 - \frac{8}{9}n_1 n_2) \tag{16}$$

and

$$\frac{\partial \phi_3}{\partial z} = \alpha n_2 - \beta n_3 - \frac{\partial \phi_1}{\partial z}, \tag{17}$$

where K_{18} is the rate coefficient in the forward direction of the charge exchange reaction,

$$H(^2S, \omega = 2) + 0^+(^4S, \omega = 4) \rightleftarrows H^+(\omega = 1) + 0(^3P, \omega = 9), \tag{18}$$

and n_4 is concentration of hydrogen atoms. The factor 8/9 in (16) is the ratio of the products of the statistical weights, ω, on the left and right hand sides of (18). Vertical drift of the ionization has been ignored in (16) and (17). It is assumed that the oxygen ions are destroyed principally by the reaction,

$$0^+ + N_2 \rightarrow NO^+ + N, \tag{19}$$

and so in (17)

$$\beta = K_{19}\, n(N_2), \tag{20}$$

where K_{19} is the rate coefficient of (19) and $n(N_2)$ is the molecular nitrogen concentration. α is the rate coefficient of the reaction

$$O + h\nu \rightarrow O^+ + e. \tag{21}$$

Photoionization and radiative recombination of atomic hydrogen are neglected in (16).

In the subsequent calculations the following expressions were adopted for the various atmospheric parameters appearing in the diffusion equations:

$$n_2(z) = n_2(400 \text{ km}) \exp(-16.8X) \text{cm}^{-3}, \tag{22}$$

$$n_4(z) = n_4(400 \text{ km}) \exp(-1.05X) \text{cm}^{-3}, \tag{23}$$

and

$$\beta(z) = \beta(400 \text{ km}) \exp(-29.4X) \text{sec}^{-1}, \tag{24}$$

where

$$X = (6770/6370+z)(z-400)/T \tag{25}$$

and

$$H_1 = (6370+z/6770)^2 T/1.05 \text{ km}, \tag{26}$$

$$H_3 = H_1/16 \text{ km}, \tag{27}$$

$$b_{12} = 3.7 \times 10^{19} (T/1000)^{1/2} \text{ cm}^{-1} \text{ sec}^{-1}, \tag{28}$$

$$b_{13} = 3.0_5 \times 10^{15} (T/1000)^{5/2} \text{ cm}^{-1} \text{ sec}^{-1}, \tag{29}$$

and

$$b_{23} = 5.7_5 \times 10^{18} (T/1000)^{1/2} \text{ cm}^{-1} \text{ sec}^{-1}. \tag{30}$$

The altitude z is measured in km. It may be noted that the altitude dependence of gravity is taken into account in equations (22) to (26). In correspondence

with (19), $\beta(z)$ was taken to be proportional to the concentration of molecular nitrogen. The collision coefficients b_{12} and b_{23} are those adopted by Dalgarno (1961), except that b_{23} has been taken to be a factor of 2.5 larger than that given by him (Dalgarno, private communication; Knof, Mason and Vanderslice, 1963). The value of b_{13} is that adopted by Bates and Patterson (1961).

3. MAINTENANCE OF THE NIGHTTIME F-REGION USING H^+ IONS

3.1 The Nighttime F-Region Solution

As the F2 layer decays after sunset, the oxygen ion concentration at high altitudes is reduced and the layer of hydrogen ions, which effectively "floats" on the oxygen ions, tends to be lowered into the region where charge exchange with atomic oxygen, as given by (18), takes place more effectively. This process replenishes the oxygen ion supply, and a steady state might be approached where enough hydrogen ions are converted to oxygen ions to balance most of the losses taking place by (19) near the F2 peak.

To obtain a solution to the problem, it is necessary to solve (13), (15), (16), and (17) simultaneously. (α is zero for the nighttime solution.) The steady-state solution of these equations is straight-forward if at any level in the ionosphere the values of n_1, n_3, ϕ_1, and ϕ_3 are known. The boundary conditions, however, cannot be stated in such a simple form, and in practice the problem must be treated as an eigenvalue problem, the values of n_1, n_3, ϕ_1, and ϕ_3 being adjusted at some level until a solution is obtained having the proper physical characteristics. These characteristics are that, (a) ϕ_3 should approach zero at high altitudes and zero at low altitudes, (b) ϕ_1 should approach zero at low altitudes and a constant value at high altitudes, (c) n_1 and n_3 should decrease monotonically with altitude at low altitudes.

The first step in the solution is to choose a starting altitude, z_o, high enough to lie above the F2 peak and low enough that O^+ is still the major ionic

constituent. For a given value of $n_3(z_o)$ a unique value for $\phi_3(z_o)$ is obtained by integrating the equations downwards and adjusting $\phi_3(z_o)$ until the lower boundary condition on n_3 is satisfied.

The second step takes into account the presence of hydrogen ions. Values of $n_1(z_o)$ and $\phi_1(z_o)$ are somewhat arbitrarily chosen, and the equations are integrated upwards; the value of $\phi_3(z_o)$ used for the upwards integration is determined by keeping the total ion flux $\phi_1(z_o) + \phi_3(z_o)$ equal to the value obtained for $\phi_3(z_o)$ in the first step. The value of $n_1(z_o)$ is then adjusted until the upwards integration solution leads to the condition $\phi_3(1500 \text{ km}) << \phi_3(z_o)$.

The third step is to again integrate downwards from z_o using the values of $\phi_3(z_o)$ and $n_1(z_o)$ obtained from the second step. If the n_1 distribution at low altitudes does not approach its chemical equilibrium value, $\phi_1(z_o)$ is adjusted and steps two and three are repeated.

This iterative technique finally leads to n_1 and n_3 distributions satisfying the correct boundary conditions. The integrations were carried out using a CDC 1604 computer, the iterations being performed automatically. The solutions at high altitudes obtained by this technique are similar to those given by Hanson and Patterson (1963), except that the present solutions are more realistic since they include the region near the F2 peak and ϕ_3 is treated as a variable instead of being taken equal to zero.

Some typical solutions obtained are plotted in Figure 1. The pertinent parameters which distinguish the solutions are given in Table 1. It is clear from the data that a downward flux of hydrogen ions of the order of 10^8 ions cm^{-2} sec^{-1} is required to maintain the F region at a stationary ionization level of about 10^5 ions cm^{-3}. This statement is valid for a rather wide range of K_{18} and β.

One of the prerequisites for the source of the (winter) nighttime F

region is that it must be able to maintain the region for several hours at a

time. An estimate of the number of H^+ ions available per cm^2 above 1000 km

can be obtained using the whistler data of Smith (1961) as an equatorial bound-

ary condition. These data were integrated downward along (dipole) magnetic

field lines to 1000 km using the diffusive equilibrium expression of Johnson

(1960). The presence of other ions was ignored. The value of n_1(1000 km) and

the total flux tube content per cm^2 derived in this manner are shown versus geo-

magnetic latitude in Figure 2 for plasma temperatures of 1000°K, 1500°K, and

2000°K. The total content is seen to increase with increasing

latitude and to be of the order of 10^{13} H^+ ions per cm^2, i.e., not too different

from that given by Hanson and Ortenberger (1961). Hence, the H^+ ion content of

the magnetosphere except near the equator appears to be sufficient to supply a

flow of 10^8 H^+ ions per cm^2 which could maintain N_mF2 at about 10^5 ions cm^{-3} for

a period of at least six hours.

3.2 The Daytime F-Region Solution

A further constraint on the hypothesis that H^+ ions are used up to maintain

the nighttime F layer is that the H^+ ions removed at night must somehow be re-

plenished. One obvious method is for them to diffuse back up into the magneto-

sphere during the daytime using the inverse charge-exchange process (18) as a

source. A calculation of this return flux was made previously by Hanson and

Patterson (1963), but there the flux of O^+ ions was arbitrarily set equal to

zero. In this section a self-consistent solution, similar to the previously

developed nighttime solution, will be obtained.

A steady-state solution is sought, and again (13), (15), (16), and (17)

must be solved simultaneously. Magnetic spreading and vertical ionization drift

velocities are ignored. The primary difference from the nighttime solution is

that the photoionization rate constant for oxygen must be taken into account.
Since the photoionization of hydrogen is neglected in this solution, it is
necessary that at some altitude z_s the flux of oxygen ions ϕ_3 must be zero if
a large upward flux of H^+ ions is to occur. There is essentially a sink
for O^+ ions at both high and low altitude, the high altitude sink resulting
from charge exchange and the low altitude sink from recombination. Thus above
z_s, ϕ_3 is positive and below ϕ_3 is negative. The positive flux of oxygen ions
is converted to H^+ ions above z_s and the upper bound on ϕ_3 is that it should
approach zero at great altitudes.

The first step in the solution of the equations is to choose a level z_s
well above the F2 peak and set $\phi_3 = 0$ at this level. The presence of H^+ ions
is at first ignored and (15) and (17) are integrated downwards choosing an
arbitrary value for $n_3(z_s)$. If the solution fails to satisfy the lower boundary
condition that $n_3 \to 0$ and $\phi_3 \to 0$ as $z \to 0$, then $n_3(z_s)$ is adjusted and the pro-
cedure repeated until a satisfactory solution is found. The final value found
for $n_3(z_s)$ is kept fixed for the remainder of the solution.

The second step is to integrate all four equations (including H^+ ions)
upwards using $n_3(z_s)$ obtained in step 1 and choosing a small positive arbitrary
value for $\phi_1(z_s)$. The value of $n_1(z_s)$ is adjusted (always keeping it less than
the H^+ ion chemical-equilibrium value to insure an upward flux of protons) in
a systematic way until a satisfactory upward solution is obtained ($\phi_3 \to 0$ as $z \to \infty$).
If z_s has been chosen too low, no such solution is possible; z_s must then be
increased and the entire solution recommenced.

The third step is to integrate the four equations downwards using the
determined values of $n_1(z_s)$, $n_3(z_s)$, $\phi_1(z_s)$, $\phi_3(z_s)$. If n_1 fails to approach
its chemical equilibrium value at low altitudes $\phi_1(z_s)$ is adjusted and steps two
and three are repeated until the lower boundary condition is satisfied. If such

a complete solution is obtained with the original z_s, then smaller values
of z_s are tried until eventually for some z_s no solution can be found. The lower
the value of z_s the greater will be the upward H^+ ion flux into the magneto-
sphere; the concentration of H^+ ions at high altitudes decreases with decreas-
ing z_s, and conversely.

Two complete solutions are shown in Figs.3 and 4. The pertinent para-
meters for the curves plotted are presented in Table 2. Curves A of Fig. 3
result from choosing $\phi_3 = 0$ at 900 km; the resultant upward flux of H^+ ions
is 5.1×10^6 cm^{-2} sec^{-1}. Curves B of Fig. 3 are obtained when ϕ_3 is set equal
to zero at 800 km. This results in an increase of ϕ_1 to 1.2×10^7 cm^{-2} sec^{-1},
but gives a smaller value of n_1 at high altitudes. An attempt to increase the
upward flux further by choosing a lower altitude (say 750 km) at which $\phi_3 = 0$
yields negative values of n_1 at high altitudes, i.e., it does not result in a
physically reasonable solution.

3.3 The Effect of Unequal Ion and Electron Temperatures

For all of the solutions thus far presented it has been assumed that the
electron, ion, and neutral particle temperatures were equal. There is con-
siderable evidence, however, that the daytime electron temperature may be con-
siderably higher than the neutral-particle temperature. Such a temperature
difference was predicted theoretically in the altitude range from 150 to 300 km
(Hanson and Johnson, 1961; Hanson, 1963; Dalgarno, McElroy, and Moffett, 1963).
Experimentally it is observed that the temperature difference on many occasions
may extend to very much greater altitudes (Nagy, Brace, Carignan and Kanal, 1963).
Though the electron temperature does not directly affect the ion-ion diffusion
coefficients, it does affect the ion scale heights and hence influences the ion
distributions. (It is a curious fact that for diffusive equilibrium the altitude
at which any two ion concentrations are equal is independent of T_e. For the case

when $T_e > T$, the 0^+ ion effective scale height increases and the H^+ ion effective scale height decreases with the net result that the level of equality remains unchanged.)

Some solutions were obtained to the daytime problem with $T_e = 2T$, and with both temperatures independent of altitude. This is somewhat unrealistic, since Dalgarno (1963) and Hanson (1963) have snown that at high altitudes the ion temperature, T_i, should approach that of the electrons. The solution is of interest however, if only to show the effect of a high electron temperature on the shape of the F2 peak. Such a solution is plotted as curve B in Figure 4 together with a solution, curve A, for which $T_e = T$ but which otherwise has the same conditions. The atmospheric parameters are given in Table 2. The effect near the F2 peak of the increased T_e is to raise h_mF2 slightly and to lower N_mF2 slightly, as might have been predicted; these differences are imperceptible on the scale of Figure 4. The total number of oxygen ions in the layer with the higher T_e is approximately ten percent larger than the one with $T_e = T_i$. Thus the collapse of T_e to T_i which may happen during an eclipse of the Sun, could lead to an increase of N_mF2 by about ten percent if the temperature equilibriation takes place in a time short compared to the recombination time of the F-region. The maximum upward flux of H^+ ions in the two cases is not greatly different, but is somewhat smaller for the larger value of T_e.

The effect of unequal ion and gas temperatures was investigated by obtaining a solution with the rather artificial assumption that $T_i = T_e = 2T$. Curves D of Figure 4 show the results.

3.4   Conclusions from Daytime F-Region Calculations

In general it can be seen from the daytime solutions presented (Table 2) that, unless the hydrogen concentration is considerably larger than generally

assumed (e.g., Kockarts and Nicolet, 1963), it is rather unlikely that the source of H^+ ions required to maintain the nighttime ionosphere can be generated in the daytime. While the presence of helium ions has been specifically ignored in this calculation, it seems improbable that their presence would invalidate the above conclusion. There is some evidence that in fact the daytime hydrogen concentrations are larger than predicted by Kockarts and Nicolet (1963) by a factor of approximately five (Hanson, Patterson, and Degaonkar, 1963). Part of this increase may be due to the fact that most of the hydrogen is evaporated from the sunlit side of the earth (Hanson and Patterson, 1963). Curves C in Figure 4 result from choosing the neutral hydrogen concentration five times larger than for curves A and B and taking $T_e = 2T$.

4.   ALTERNATIVES TO DIFFUSIVE REPLENISHMENT OF THE DAYTIME F-REGION

4.1 Photoionization of Atomic Hydrogen

The daytime protonosphere might concievably be replenished by direct photoionization of atomic hydrogen. An estimate will now be made of the importance of this source. If the neutral hydrogen were distributed hydrostatically with altitude, the production rate, S_H, of hydrogen ions per cm^2 column above 400 km would be given by,

$$S_H = \alpha_H \, n_4(400 \text{ km}) \, H_1 (1 + \frac{2H_1}{r}) \qquad (31)$$

where α_H is the photoionization rate coefficient of atomic hydrogen, n_4 (400 km) is the neutral hydrogen concentration at 400 km, r is the distance of the 400 km level from the center of the earth, and H_1 is the neutral hydrogen scale height at 400 km. This equation should give an upper limit to the magnitude of the source of hydrogen ions in so far that the actual neutral hydrogen distribution in the exosphere falls below the hydrostatic distribution. Taking n_4 (400 km) = 10^5 cm^{-3}, T = 1000°K, H_1 = 1000 km and $\alpha_H = 4.5 \times 10^{-7} \text{ sec}^{-1}$ (Hinteregger, 1960)

gives $S_H = 5.8 \times 10^6$ cm^{-2} sec^{-1}. The number produced during a 12 hour day would thus be 2×10^{11} cm^{-2} which is too small to significantly affect the F region. The source could be significant however, to the establishment of the protonosphere itself.

4.2 Local Ionization Source

Near the auroral zone the precipitating energetic particles certainly constitute an appreciable source of nighttime ionization. It is difficult to show conclusively that the ionization caused by energetic particles at lower latitudes is negligible, but there are certain constraints which this particle flux must satisfy. It must cause more than 10^8 ionization events per cm^2 per sec above the F2 peak, and it must do this without raising the electron temperature more than a few hundred degrees above the gas temperature. Further, it must not, nor should the secondary electrons, excite more oxygen atoms to the 'D state than it ionizes. The number of oxygen atoms raised to the 'S state must be considerably less than the number of ionizing events. To cause the requisite ionization in the proper altitude range a flux of fast electrons greater than 10^9 cm^{-2} sec^{-1} would be required.

It seems doubtful that a particle flux of the correct pitch angle and energy distribution could be constructed, and even less likely it would exist on such a regular basis. However, it is quite possible that these particle fluxes could play a significant role on some occasions.

4.3 The Effect of Vertical Drift

An alternative hypothesis to maintaining the nighttime F layer by charge exchange is that the effective decay rate of the layer may be decreased in an entirely different manner. While it is possible that the rate coefficient K_{19} may have a strong positive temperature dependence, there is no evidence that this is so. On the other hand it can be shown that an upward drift velocity w of

the ionization tends to raise the F2 peak, and to decrease the effective decay
rate of the layer (Martyn, 1947; Martyn, 1956; Duncan, 1956). The changes in
the height of the F2 peak by this process have been calculated for the electric
field associated the S_q current system (Hirono, 1955). There is some uncertainty
as to the exact phase of this drift velocity, but there is general agreement
that the drift is upwards between midnight and sunrise, which is the approximate
time interval when the decay rate appears to have its smallest value. In order
to make the subsequent analysis tractable it will be explicitly assumed that
the vertical ionization drift velocity is independent of altitude. This is an
important restriction and many of the conclusions will require modification if
w varies greatly with altitude.

When oxygen ions alone are present (15) reduces to the simpler form,

$$\phi_3 = -\frac{2b_{23}}{n_2}\left(\frac{\partial n_3}{\partial z} + \frac{n_3}{2H}\right) , \tag{32}$$

and (12) gives directly

$$\frac{\partial n_3}{\partial t} = -\beta n_3 - \frac{\partial \phi_3}{\partial z} - w\frac{\partial n_3}{\partial z} . \tag{33}$$

H denotes the scale height of atomic oxygen given by (27) (the subscript 3 has
been dropped for convenience).

The solution of (32) and (33) has been discussed by Martyn (1956), Duncan
(1956) and Dungey (1956) when H is independent of altitude. Martyn (1956) show-
ed that if $w = 0$ and $\beta \propto e^{-z/H}$, a stable layer of ions would form having the
shape of a Chapman layer. He further showed that the layer would decay exponential-
ly with time and have a time constant $1/\lambda$, where λ is the value of the loss coef-
ficient at the F2 peak and is given by $D/4H^2$. D is the familiar ambipolar dif-

fusion coefficient, given approximately by

$$D = 2b_{23}/n_2(z). \tag{34}$$

Duncan (1956) extended Martyn's work with $w(z)$ = constant and showed that a stable Chapman layer would still be formed but that it would be displaced vertically from its normal position by the amount $\Delta Z = H$ arc sinh wH/D, where D is determined at the F2 peak when $w = 0$. Dungey (1956) showed that so long as β decreased exponentially with altitude, a stable layer would form which decayed uniformly and exponentially with time, but that the layer was not necessarily Chapman-like. Numerical solutions to this problem (including photoionization) were presented by Rishbeth and Barron (1960) for various scale height of β; they also took account of vertical gradients in the atmospheric temperature.

Particular solutions to (32) and (33) can be expressed as

$$n_3(z,t) = N_3(z)e^{-\lambda t} \tag{35}$$

and

$$\phi_3(z,t) = \overline{\Phi}_3(z)e^{-\lambda t}, \tag{36}$$

where N_3, $\overline{\Phi}_3$ and λ are independent of time. Substituting (35) and (36) into (32) and (33) yields the equations,

$$\frac{\partial N_3}{\partial z} = -\frac{N_3}{2H} - \frac{n_2\overline{\Phi}_3}{2b_{23}} \tag{37}$$

and

$$\frac{\partial \overline{\Phi}_3}{\partial z} = (\lambda - \beta)N_3 - w\frac{\partial N_3}{\partial z}, \tag{38}$$

where it can be seen that λ is the value of the loss coefficient at the altitude where $\overline{\Phi}_3 + N_3w$ is maximal; in general this will not occur where N_3 is maximal.

Numerical solutions to (37) and (38) are obtained in the following manner. The equations are solved, for a particular choice of n_2 (400 km), β(400 km) and T in (22) and (24), in the region below a reference altitude z_0 which is choosen

sufficiently high that O^+ can be assumed to be in diffusive equilibrium. The upper boundary condition on Φ_3,

$$\Phi_3(z_o) = - \int_{z_o}^{\infty} (\lambda - \beta) N_3 \, dz - w N_3(z_o), \tag{39}$$

can then be evaluated to give

$$\Phi_3(z_o) = 2 H N_3(z_o) \left\{ \frac{2}{9} \beta (400 \text{ km}) \exp[7(400 - z_o)/4H] - \lambda - w/2H \right\}. \tag{40}$$

It may be noted that β has the scale height of molecular nitrogen, $\frac{4}{7} H$. Eq. (40) expresses the interdependence of $\Phi_3(z_o)$ and λ.

The equations are linear in N_3 and so $N_3(z_o)$ may be arbitrarily assigned. An arbitrary value for $\Phi_3(z_o)$ is chosen and a corresponding value of λ derived from (40). Eqs. (36) and (37) are then integrated downwards to determine if the solution satisfies the lower boundary conditions that $\Phi_3 \to 0$ and $N_3 \to 0$ as $z \to 0$ (the method of Rishbeth and Barron, 1960). If the solution diverges as $z \to 0$, $\Phi_3(z_o)$ is adjusted, a new value for λ is derived from (40) and the procedure is repeated until a satisfactory solution is found.

To estimate the possible vertical drift of the nighttime F layer a particular set of ionosonde records (kindly supplied to us by G. H. Stonehocker of the National Bureau of Standards) will be examined. These were obtained from San Salvador (magnetic dip $\simeq 55°$) in the early hours of 12 April 1961; the ionosonde data were confirmed by simultaneous measurements from a Blue Scout Rocket (Sagalyn and Smiddy, 1964).

The neutral particle temperature derived from the radio-noise flux for this time (Harris and Priester, 1962) is about 720°K. This temperature is adopted to determine the pertinent parameters for the ensuing calculations. The oxygen atom concentration at the height of the F2 peak (? 30 km) is only $9.4 \times 10^7 \text{ cm}^{-3}$ for $T = 720°K$, according to Harris and Priester (1962). It is possible to obtain

a good fit to the measurements for various combinations of β and vertical drift velocities w. With w = 0, and the collision coefficient b_{23} = 4.88 x 10^{18} cm^{-1} sec^{-1}, it is found that β(330) = 7.75 x 10^{-4} sec^{-1} yields a best fit. In this case, however, λ = 1.16 x 10^{-3} sec^{-1}, so that the layer would be almost completely dissipated in less than one hour. The ionosonde data, on the other hand, show that N_mF2 was nearly constant. If it is assumed that (19) is the appropriate loss process, and Harris and Priester's (1962) value for the molecular nitrogen concentration of 1 x 10^7 cm^{-3} at 330 km is accepted, then K_{19} = 7.75 x 10^{-11} cm^3 sec^{-1} for β(330) = 7.75 x 10^{-4} sec^{-1}. Such a large value for K_{19} would clearly be difficult to reconcile with the daytime F region (Bates and Nicolet, 1960).

Smaller values of β(330 km) and of λ can be obtained if a positive value is adopted for w. The dependence of h_mF2 on w can be estimated in the following manner. From (32)

$$\left(\frac{\partial N_3}{\partial z} \right)_{h_m F2} = 0 = \left[- \frac{N_3}{2H} - \frac{\Phi_T}{D} + \frac{N_3 w}{D} \right]_{h_m F2} \tag{41}$$

where $\Phi_T = \Phi_3 + N_3 w$ is the sum of the fluxes due to diffusion and vertical drift. Hence, when w = 0,

$$W = - \left[\Phi_T (w=0)/N_3 \right]_{h_m F2} = \frac{D^o}{2H} , \tag{42}$$

where W and D^o are respectively the downward diffusion velocity of O^+ ions and the ambipolar diffusion coefficient as defined by (33) at the F2 peak when w = 0. If the effect of w were to displace the F region vertically with no distortion of shape, the flux through the F2 peak should decrease as $\beta(h_m F2)$. Hence, for

finite w,

$$\Phi_T(h_mF2) = \Phi_T(w=0) \exp(-7\Delta Z/4H) \qquad (43)$$

and

$$D(h_mF2) = D^\circ \exp(\Delta Z/H), \qquad (44)$$

where ΔZ represents the vertical displacement of the F2 peak from its position when w = 0 due to the drift velocity w. ΔZ has, of course, the same sign as w. Substituting (43) and (44) into (41), it follows that

$$w = W[\exp(\Delta Z/H) - \exp(-7\Delta Z/4H)]. \qquad (45)$$

If β had been taken to have the same scale height as atomic oxygen, (44) would reduce to $(w/W) = 2 \sinh (\Delta Z/H)$, as deduced by Duncan (1956).

From the numerical solution of (37) and (38), it is found that there is some shape distortion of the F region when β has a scale height of $\frac{4}{7}$ H, with a tendency for the region below the F2 peak to become thicker as w increases. This distortion causes the effective β of the layer (and therefore $\Phi_T(h_mF2)$) to fall off more slowly with altitude than the concentration of molecular nitrogen. Empirically, the expression

$$w = W[\exp(\Delta Z/H) - \exp(-5\Delta Z/4H)] \qquad (46)$$

is found to provide an excellent fit to the numerical calculations for both positive and negative values of w as is shown in Figure 5. The parameters associated with this figure are shown in Table 3. For small $(\Delta Z/H)$ the expression (46) can be written

$$\Delta Z = \frac{4}{9} H \frac{w}{W} . \tag{47}$$

The work of Dungey (1956) implies that when $w = 0$ and when H is assumed to be independent of altitude the value of $\beta H^2/D$ at the F2 peak depends only on the scale heights of β and D. For the scale heights chosen here, the numerical solution of (37) and (38), with the correct altitude dependence of H included, shows that for $w = 0$,

$$[\beta H^2/D^\circ]_{h_m F2} = 0.13. \tag{48}$$

Eq. (48) allows $h_m F2$ to be determined when $w = 0$ if T is given and β and D are known at one altitude. Thus, (46) together with (48) enable $h_m F2$ to be calculated for any value of w. Such an analysis has been carried out for comparison with the ionosonde data; the results are shown in Figures 6 and 7. In Figure 6 the collision coefficient for the diffusion of O^+ through O has been taken to be $b_{23} = 4.88 \times 10^{18}$ cm^{-1} sec^{-1} (Mason and Vanderslice, 1964; Dalgarno, private communication), whereas in Figure 7 the value previously suggested by Dalgarno (1961) $b_{23} = 1.96 \times 10^{18}$ cm^{-1} sec^{-1} has been used. Alternatively, the concentration of O may be considered two and one half times larger for Figure 7 than for Figure 6, but with the larger collision coefficient in both cases. The factor $\sin^2 I \approx 2/3$ (San Salvador), which should properly be used to modify the collision coefficients, has not been included in the above calculations.

It is apparent from Figure 6 that if $\beta(h_m F2)$ is to be less than 10^{-4} sec^{-1}, then w must be greater than 100 m/sec. The value of $h_m F2$ with $w = 0$ would then be less than 300 km. The magnitude of λ, which actually determines the decay rate of the layer, is larger than $\beta(h_m F2)$ and depends on w. The expression

$$\lambda/\beta(h_m F2) = 1.5 + 0.2(w/W) \tag{49}$$

has been obtained from the numerical calculations and is accurate to better than 10 percent for $|w/\tilde{w}| < 4$. The ionosonde data from San Salvador indicate that in fact the decay constant for $N_m F2$ was approximately 3×10^{-5} sec^{-1} during the period from 1:00 AM to 3:00 AM local time ($h_m F2$ remained nearly constant over this time interval). Even if the smaller diffusion coefficient is taken, Figure 7 shows that $w > 31$ m/sec for $\beta(h_m F2) < 10^{-4}$ sec^{-1}. The height of the F2 peak with $w = 0$ would be less than 315 km for this case. It should be remarked that the thermal contraction of the ionosphere at night will give rise to an effective source of ionization which could be taken into account by adding a term $\beta_T N_3$ to the right hand side of (38). The magnitude of β_T is approximately 10^{-5} sec^{-1}. This term should properly be dealt with when λ is of this order, but has not been considered in these calculations.

One of the explicit assumptions in the above discussion is that the scale height of the ambipolar diffusion coefficient is the same as that of neutral atomic oxygen. This is a good assumption provided that atomic oxygen is the predominant neutral constituent. Below the altitude where the ratio $n(0)/n(N_2) = 1$, this assumption is no longer a good approximation; should this equality level be near the F2 peak, then the above simple analysis would need to be modified in a straightforward, but not trivial, manner. The conclusions would be altered somewhat, but qualitatively would remain the same.

The quantity a/H, where a is the distance measured downwards from the F2 peak to the level where the ion concentration is $\frac{1}{2} N_m F2$, may be obtained from the numerical solutions. The value of a/H is not too sensitive to w, and for $w = 0$ a/H = 1.18. For a Chapman layer a/H = 1.46. Nisbet and Quinn (1963) have deduced H from some ionograms by fitting a Chapman layer to the ion concentration below the F2 peak. To take account of the different form of the recombination

coefficient for a Chapman layer, the present results would indicate that their scale heights should be increased by a factor of 1.46/1.18 which would bring about rather good agreement between their derived scale heights and the atmospheric models they discuss.

It would appear from the above analysis of the ionosonde data that a vertical ionization drift velocity of at least 30 m/sec must have been present for a period greater than two hours. The expression for the vertical drift velocity near the F2 peak is closely approximated by (Dougherty 1961)

$$w = \frac{E_y}{B} \cos I + U \sin I \cos I, \tag{50}$$

where E_y is the eastward component of the electric field, I is the magnetic dip angle and U is the southward component of the atmospheric wind velocity. Taking $E_y = 0$, a wind velocity $U = 64$ m/sec is required to establish the minimum value of $w = 30$ m/sec derived above. As Dougherty (1961) has emphasized, the ions exert a drag on the atmospheric wind system which tends to damp out the relative ion-neutral horizontal velocity. The time constant associated with this mechanism is

$$\tau = mb_{23}/kTn_3 \tag{51}$$

and is approximately 35 minutes and 90 minutes respectively for the smaller and larger values of b_{23} used here. As stated above, the F2 peak height remained nearly constant for more than two hours, indicating that w did not change appreciably in this time.

The neutral wind system could be maintained in spite of the ion drag forces by a horizontal atmospheric pressure gradient. The slope S of the constant pressure surface required to maintain a given w can be estimated from the expression

$$S = (H/p) \frac{\partial p}{\partial x} = (w/b_{23})n_3 H \cot I, \tag{52}$$

where p is the atmospheric pressure and x is the north-south coordinate.

For the particular case described here, (52) yields a value of 10^{-3} for S.

The existence of such a small slope certainly cannot be ruled out on the basis

of our present knowledge of the atmospheric pressure at these heights.

It is of course futile to try to draw any general conclusions from the

behavior of the F2 peak on only one occasion at one location. A cursory exam-

ination of h_mF2 data shows however that this occasion was not very exceptional.

It is clear that a closer look at the abundant ionosonde data from this point

of view may be profitable in understanding how the F region is maintained at

night, and may also contribute to our understanding of global wind patterns (and/or

ionospheric electric fields). For example, if the various atmospheric parameters

can be well enough established so that w can be reliably inferred from h_mF2 in

the manner described above, then (50) and the equation (Dougherty, 1961)

$$u_i = U \cos^2 I - \frac{E_y}{B} \sin I - \frac{D}{2H} \sin I \cos I \tag{53}$$

can be used to determine E_y and U at the F2 peak uniquely if the north-south

component of the ionization drift velocity u_i is simultaneously measured. Sol-

ution of (50) and (53) gives

$$U = w \tan I + \frac{D}{2H} \cos I + u_i \tag{54}$$

and

$$E_y/B = w \cos I - \frac{D}{2H} \sin I \cos I - u_i \sin I. \tag{55}$$

While it is true that measurements of u_i are somewhat ambiguous, those recently

reported by Rao and Rao (1963) appear to have the correct order of magnitude to

be consistent with the analysis presented here.

It should be emphasized that maintenance of a high h_mF2 by a change in β differs fundamentally from maintenance by a vertical drift of ionization. It is true that the altitude of the equilibrium $(w = 0)F2$ peak can be increased by increasing β. The net effect, however, is to increase the rate of decay of the layer. (The increase in decay rate is not large; increasing β by a factor 10 increases λ by only a factor of 2.) On the other hand, if the F2 peak height is raised by vertical drift, the net effect is to decrease the rate of decay of the layer. On many occasions h_mF2 increases at night; while this could be due to several causes, it is not impossible that the effect occurs naturally because the ionosonde is sampling a different part of ionosphere at different times due to the horizontal drift of the ionization which accompanies the vertical drift when the movement is due to electric fields (rather than neutral winds which, in effect, cause only a vertical displacement of the layer).

5. CONCLUSIONS

It seems most unlikely that the F region is maintained at night by utilizing the protonosphere as an ionization source. On the other hand, the vertical drift of the ionization to a region where the recombination is relatively slow appears to be an attractive alternative. The simultaneous observation of h_mF2 and the ionization drift velocity at night can be utilized to infer the electric field and the southward component of the atmospheric wind velocity at F region heights.

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CAPTIONS

Figure 1   The effect of changing β on the nighttime distribution with altitude of hydrogen ions and atomic oxygen ions. The calculations were carried out   with $n(0/400 \text{ km}) = 8.0 \times 10^7 \text{ cm}^{-3}$, $n(H/400 \text{ km}) = 10^4 \text{ cm}^{-3}$, $K_{18} = 4.0 \times 10^{-10} \text{ cm}^3 \text{ sec}^{-1}$ and $T = 1000°K$. Curves A correspond to $\beta(400 \text{ km}) = 10^{-5} \text{ sec}^{-1}$ while curves B correspond to $\beta(400 \text{ km}) = 10^{-6} \text{ sec}^{-1}$.

Figure 2   Whistler measurements. The full curves show the electron concentrations   at 1000 km versus geomagnetic latitude as deduced from whistler measurements using the diffusive equilibrium expression of Johnson (1960). The plasma temperature is indicated on each curve. The dashed curves show the total electron content above 1000 km in a tube of flux with a 1 cm^2 base as a function of geomagnetic latitude.

Figure 3   Daytime distribution with altitude of hydrogen ions and atomic oxygen ions. The calculations were carried out with $n(0/400 \text{ km}) = 1.9 \times 10^8 \text{ cm}^{-3}$, $n(H/400 \text{ km}) = 10^4 \text{ cm}^{-3}$, $\beta(400 \text{ km}) = 10^{-5} \text{ sec}^{-1}$, $K_{18} = 9.3 \times 10^{-10} \text{ cm}^3 \text{ sec}^{-1}$, $\alpha = 7.0 \times 10^{-7} \text{ sec}^{-1}$, and $T = 1350°K$. Curves A correspond to $\phi_3(0^+/900 \text{ km}) = 0$ while curves B correspond to $\phi_3(0^+/800 \text{ km}) = 0$.

Figure 4   Daytime distribution with altitude of hydrogen ions and atomic oxygen ions. The calculations were carried out with $n(0/400 \text{ km}) = 8.0 \times 10^7 \text{ cm}^{-3}$, $\beta(400 \text{ km}) = 2.0 \times 10^{-6} \text{ sec}^{-1}$, $K_{18} = 9.3 \times 10^{-10} \text{ cm}^3 \text{ sec}^{-1}$, $\alpha = 3.0 \times 10^{-7} \text{ sec}^{-1}$. $T = 1000°K$. The curves labelled A correspond to $n(H/400 \text{ km}) = 2.0 \times 10^4 \text{ cm}^{-3}$ and $T_e = T$. Those labelled B correspond to $n(H/400 \text{ km}) = 2.0 \times 10^4 \text{ cm}^{-3}$ and $T_e = 2T$ while those labelled C correspond to $n(H/400 \text{ km}) = 1.0 \times 10^5 \text{ cm}^{-3}$ and $T_e = 2T$. Curves D have $n(H/400 \text{ km}) = 2.0 \times 10^4 \text{ cm}^{-3}$ and $T_i = T_e = 2T$.

Figure 5   A plot of the increase in h_mF2 measured in scale heights versus drift velocity normalized by the vertical diffusion velocity when $w = 0$. The plotted points were determined from numerical solutions to (37) and (38) for the conditions shown in Table 3.

Figure 6   The relationships between β at an F2 peak height of 330 km and the drift velocity w required to maintain h_mF2 at 330 km. $h_m°F2$ is the height of the F2 peak with $w = 0$ and W is the downward vertical diffusion velocity (D/2H) at this height. For example, to maintain h_mF2 at 330 km with $\beta = 10^{-4}$ sec^{-1} at this level $h_m°F2$ is found to be 299 km, $w = 103$ m sec^{-1}, and $W = 60$ m sec^{-1}. The collision coefficient b_{23} was taken to be 4.88×10^{18} cm^{-1} sec^{-1}.

Figure 7   Similar to Figure 6 except that b_{23} is smaller by a factor of 2.5.

T A B L E 1

NIGHTTIME SOLUTIONS

| Fig. | Curves | 400 km | | | $h_m F2(km)$ | $T(°K)$ | $\phi_1(max)$ cm^{-2} sec^{-1} | K_{18} cm^3 sec^{-1} |
		$n(H)$ cm^{-3}	$n(O)$ cm^{-3}	$B(sec^{-1})$				
1	A	10^4	8.0×10^7	10^{-5}	329	1000	-3.3×10^8	4.0×10^{-10}
1	B	10^4	8.0×10^7	10^{-6}	280	1000	-2.0×10^8	4.0×10^{-10}

T A B L E 2

DAYTIME SOLUTIONS

400 km

Fig.	Curves	$n(H)$ cm^{-3}	$n(O)$ cm^{-3}	β sec^{-1}	$T(°K)$	$T_e(°K)$	z_s(km)	ϕ_1(max) cm^{-2} sec^{-1}	α(sec^{-1})	K_{18} cm^3 sec^{-1}
3	A	1.0×10^4	1.9×10^8	10^{-5}	1350	1350	900	5.1×10^6	7.0×10^{-7}	9.3×10^{-10}
3	B	1.0×10^4	1.9×10^8	10^{-5}	1350	1350	800	1.2×10^7	7.0×10^{-7}	9.3×10^{-10}
4	A	2.0×10^4	8.0×10^7	2.0×10^{-6}	1000	1000	600	6.3×10^6	3.0×10^{-7}	9.3×10^{-10}
4	B	2.0×10^4	8.0×10^7	2.0×10^{-6}	1000	2000	625	4.1×10^6	3.0×10^{-7}	9.3×10^{-10}
4	C	1.0×10^5	8.0×10^7	2.0×10^{-6}	1000	2000	600	5.5×10^6	3.0×10^{-7}	9.3×10^{-10}
4	D	2.0×10^4	8.0×10^7	2.0×10^{-6}	1000	2000*	500	1.2×10^7	3.0×10^{-7}	9.3×10^{-10}

* For this case $T_i = T_e$

T A B L E   3

PARAMETERS ASSOCIATED WITH FIGURE 5

400 km		$T \ °K$	$H \ km$	$W \ m/sec$	$h°_m F2(km)$	b_{23}
$\beta \ sec^{-1}$	$n(0) \ cm^{-3}$					$cm^{-1} \ sec^{-1}$
1×10^{-5}	1.9×10^{8}	1350	79.6	9.34	350	4.88×10^{18}
1.5×10^{-6}	2×10^{7}	720	41.8	36.9	279	1.96×10^{18}

FIGURE

FIGURE 2

FIGURE 3

ALTITUDE (km)

ION CONCENTRATION (cm⁻³)

FIGURE 4

T = 1000 °K

$$\left(\frac{w}{W}\right) = e^{\frac{\Delta z}{H}} - e^{\frac{-5\Delta z}{4H}}$$
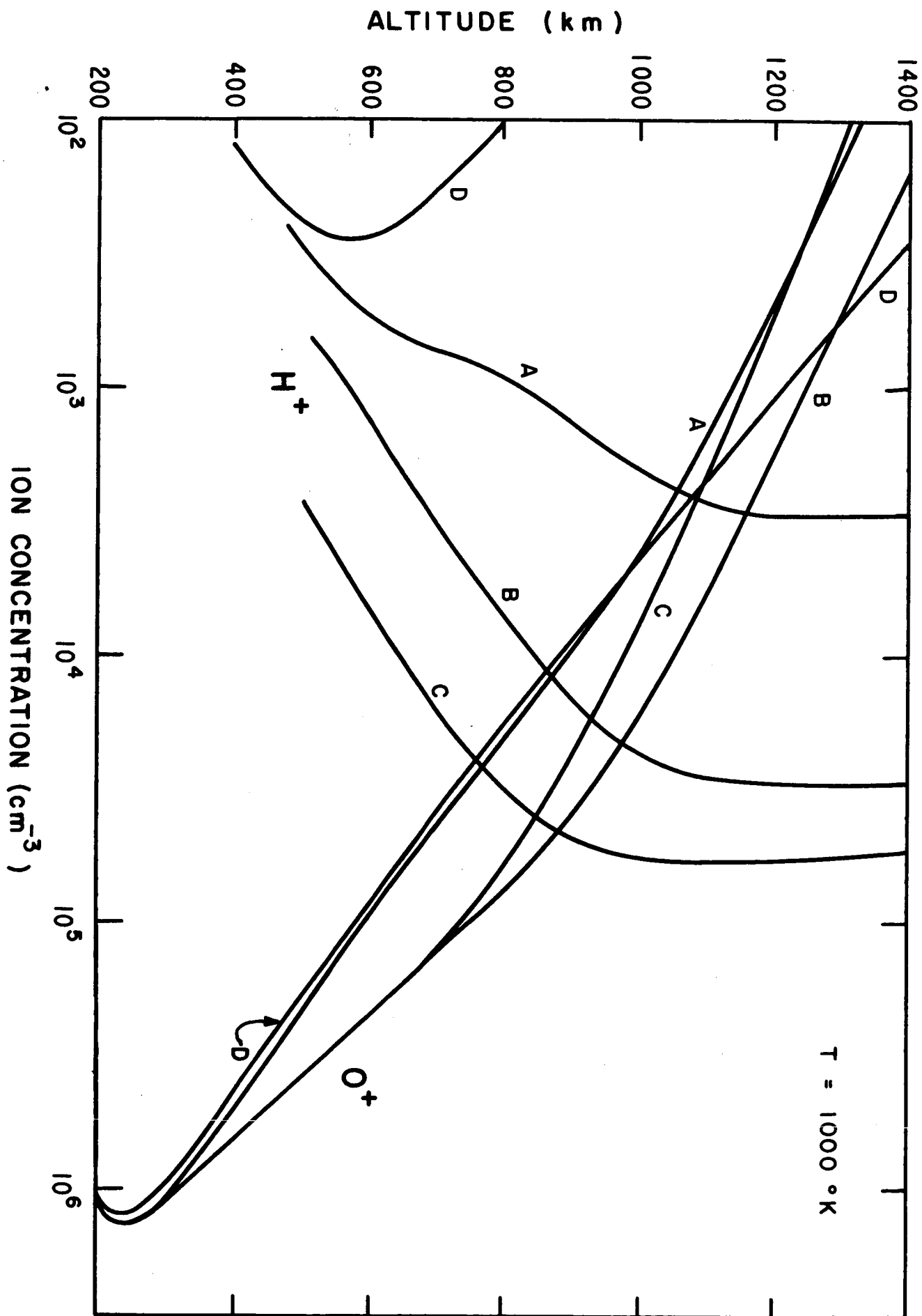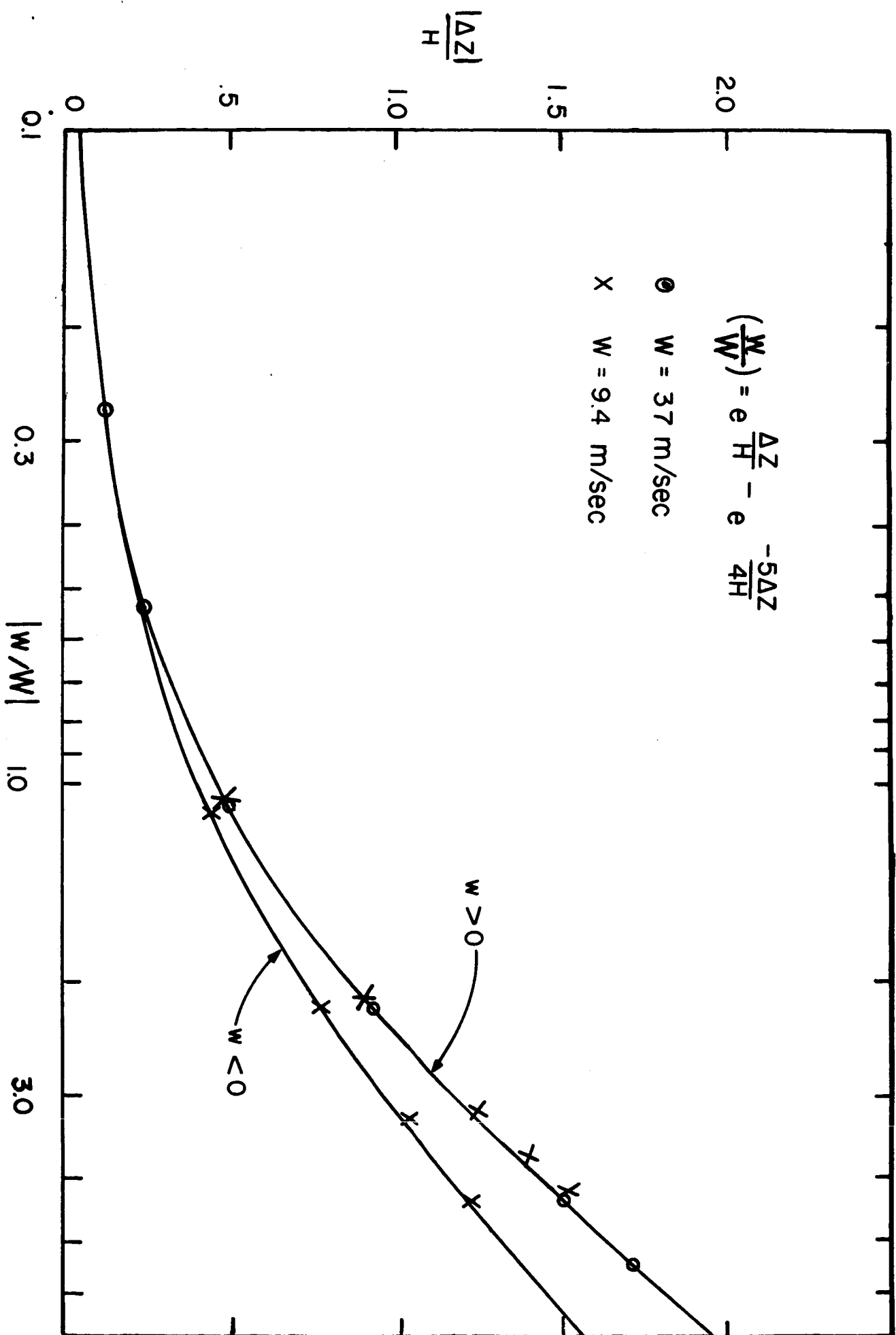
⊙ W = 37 m/sec

✕ W = 9.4 m/sec

w > 0

w < 0

w < 0

$\frac{|\Delta z|}{H}$

$|w/W|$
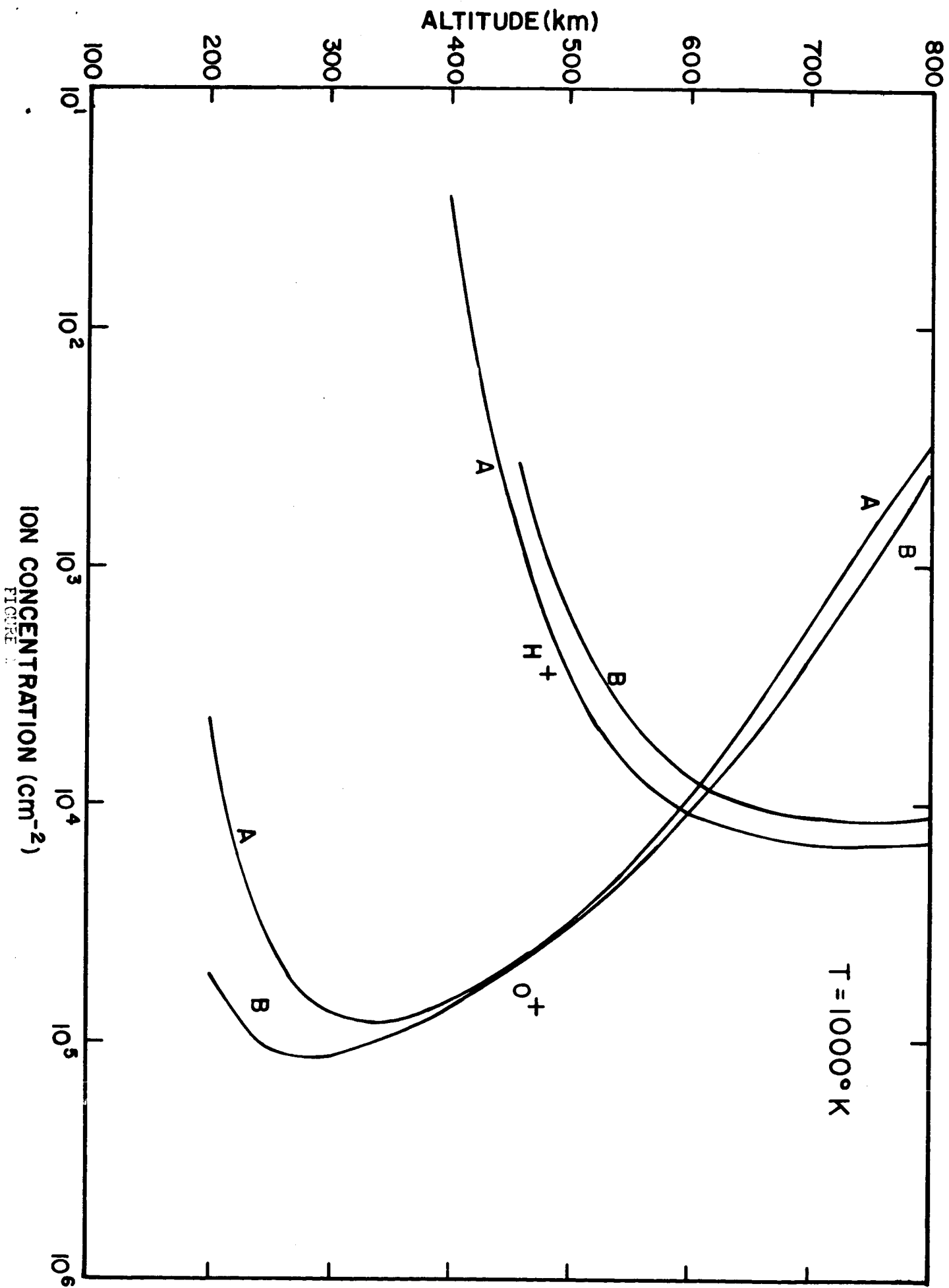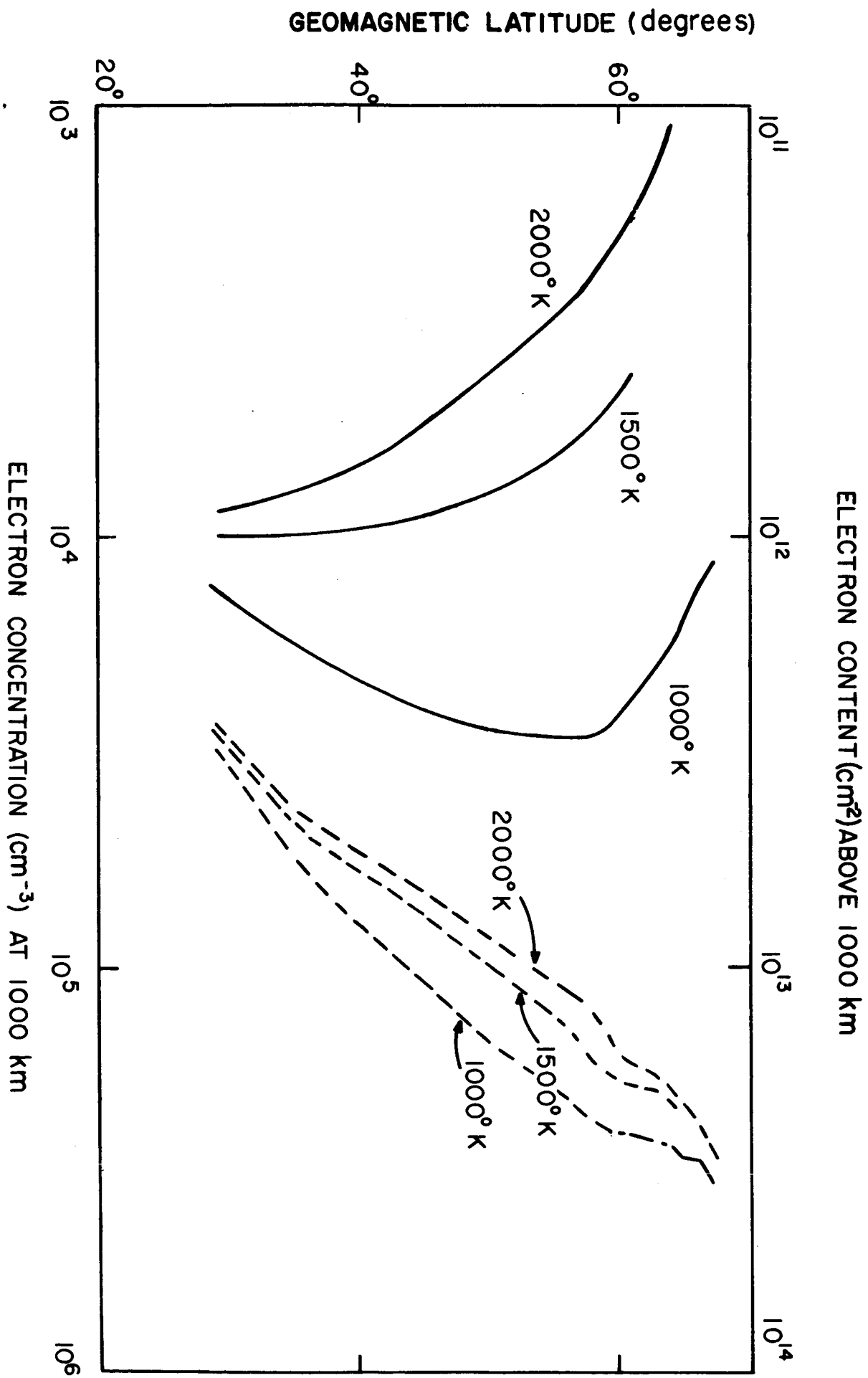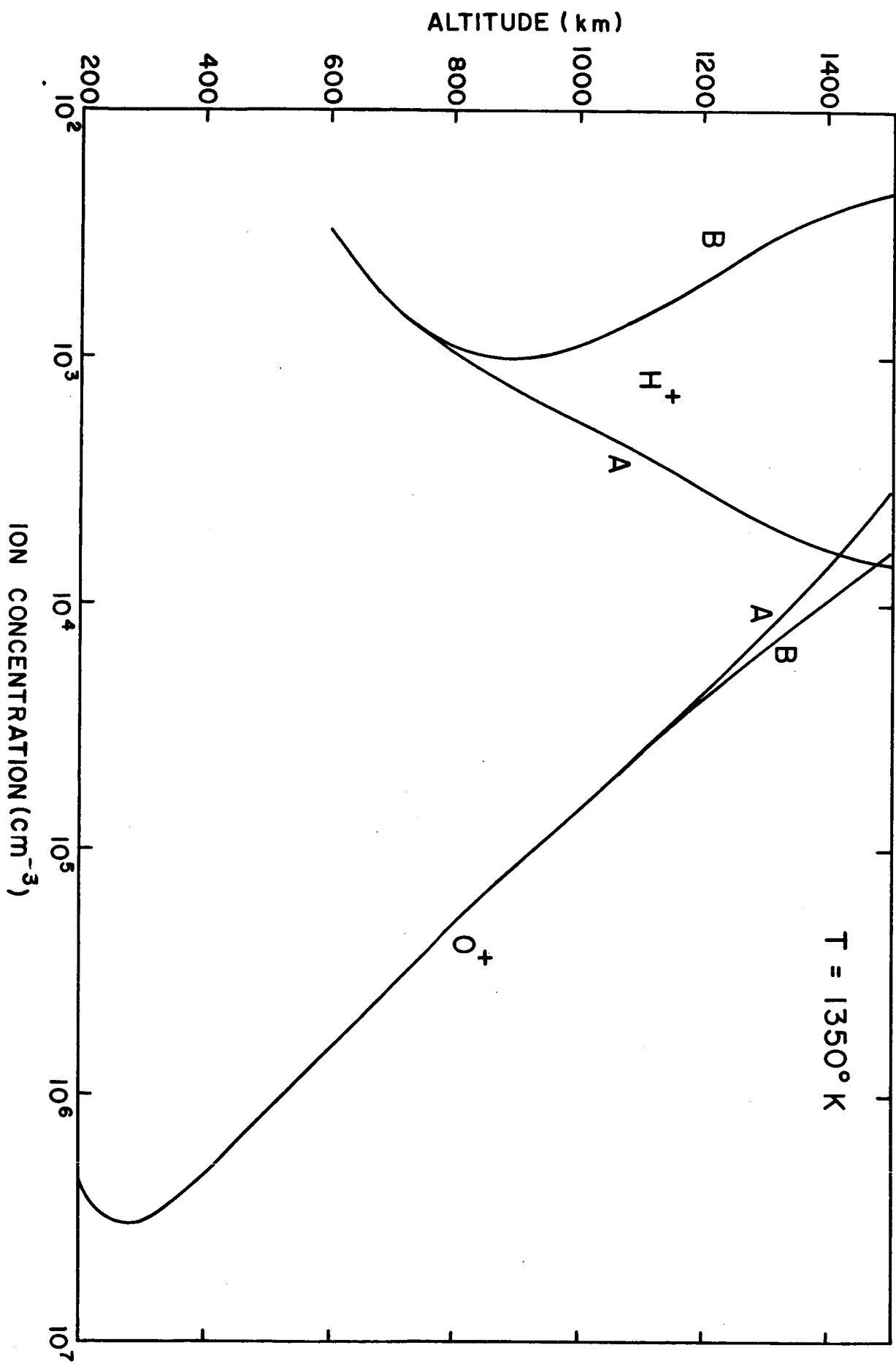
FIGURE 5

FIGURE 6

FIGURE 7